

Abyssal circulation around New Zealand—a comparison between observations and a global circulation model

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Abstract

Observations of abyssal currents off eastern New Zealand are compared to results from the Los Alamos National Laboratory (LANL) global ocean circulation model. Physical oceanographic measurements are few along the 6000-km long path of the abyssal flow so they are supplemented by geological data including bottom photographs, nephelometer profiles, sediment analyses, and high resolution seismic profiles. While greatly increasing the spatial coverage of the observations, the geological data have limitations concerning temporal aspects of the circulation, e.g., the resolution of seismic records restricts identification of bottom current action to periods of $\sim 12,000$ years or more. Despite these limitations, the model compares well with observations for the region south of Chatham Rise (43°S). There, the model shows a highly energetic, topographically steered flow that coincides with zones of seafloor erosion, active bedload transport, and prominent benthic nepheloid layers. In contrast, correlation is less clear to the north of Chatham Rise. Model output and observations agree on flow directions in regions of marked topography, but in areas of subdued relief, model current directions depart from reality. Such departures are because the model bathymetry has (1) a grid too coarse to resolve small but key current pathways, and (2) step-like contours that artificially guide the flow even though in nature the seabed may have low relief. The model also underestimates current intensity as measured by eddy kinetic energy (EKE) and volume transport. However, these deficiencies can be reduced by improving the model bathymetry and better representing the oceanic processes such as the interaction of Rossby waves with the bathymetry. © 1999 Elsevier Science B.V. All rights reserved.

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1. Introduction

The extensive submarine microcontinent off New Zealand is the Pacific gateway for the largest, deep

western boundary current (DWBC), (e.g., Warren, 1981). Furthermore, the microcontinent projects deep into the Southern Ocean thus forming one of the major constrictions of the Antarctic Circumpolar Current (ACC) (e.g., Patterson and Whitworth, 1990). Yet despite the significance of the New Zealand region, the deep oceanic circulation is not well constrained by physical oceanographic observations. Hy-

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drographic sections are scattered widely and long-term (> 0.5 year) current meter stations are even more scarce.

As a result, some reliance has been placed on global circulation models (Webb et al., 1991; Semtner and Chervin, 1992) to gain a perspective of the abyssal flow through the region. Recently, Carter and McCave (1994, 1997) and McCave and Carter (1997) reviewed sedimentary and geophysical data to evaluate the erosional and depositional regime beneath the DWBC. This database has wide geographic coverage and, when combined with physical oceanographic information, provides reasonable observational control of the path and strength of the DWBC. We are, therefore, in a position to test numerical model output for the abyssal circulation along the 6000-km long course of the DWBC between 58°S and 25°S (Fig. 1). This reach of the current is well suited for the comparison in light of strong regional differences in the flow caused by a variable bathymetry and by interaction with the ACC (Carter and McCave, 1997).

1.1. Numerical model

The simulated abyssal circulation is the 10-year mean flow derived from the Los Alamos National Laboratory (LANL) global ocean model. This has a higher resolution than its predecessor, the 0.5° global model of Semtner and Chervin (1992). The LANL model grid spacing is 0.28° longitude by $\cos(\text{latitude}) \times 0.28^{\circ}$ latitude equalling 22 km at 45°S . Twenty vertical levels are designated with spacings about 500 m apart below 2000 m depth and closer spacings in shallower water. Bathymetry is derived by setting the depth at each grid point to that of the model level nearest to the observed depth in the digital bathymetric database, ETOPO5. In profile, a sloping seafloor thus appears as a flight of steps of finite height. The model was integrated for the 10-year period 1985–1994 using three daily averages of observed winds, surface heat, and freshwater fluxes from a monthly climatology (Maltrud, submitted; McClean et al., 1997).

As this paper deals mainly with the path and energetics of the DWBC/ACC, the model circulation is presented as mean velocity vectors, and con-

tours of velocity variance. Mean velocity fields are displayed for model levels below 2000 m depth, this being the commonly used level-of-no-motion marking the upper limit of the abyssal circulation (e.g., Warren, 1981). In addition, the flow near the ocean floor is presented to highlight topographic steering effects.

The shallower flow is examined off southernmost New Zealand to determine the path of the wind-driven ACC which is sufficiently deep-reaching to influence the underlying DWBC (Gordon, 1975).

1.2. Observational database

An important source of information on abyssal sedimentation is the data reports and associated publications from the Australasian–Antarctic research cruises of the *USNS Eltanin* (Fig. 2) which are summarised in Carter and McCave (1994, 1997). Additional information comes from recent studies of DWBC sedimentation off Campbell Plateau (Carter and McCave, 1997), Bounty Trough (Carter and Mitchell, 1987; Carter et al., 1990; Carter and Carter, 1996) and between Chatham Rise and Kermadec Trench (Carter and McCave, 1994; McCave and Carter, 1997). These regional synopses rely on sediment analyses, nephelometer profiles and bottom photographs, together with airgun and 3.5 kHz seismic profiles.

As a proxy for modern currents, geological data have some limitations regarding the timing and quantification of causal hydraulic processes. Perhaps the most reliable information comes from bottom photographs. Through interpretation of sediment type, bedforms, benthic organisms and other features, a semi-quantitative assessment of the bottom flow can be made, e.g., McCave and Tucholke (1986). However, interpretations from photographs can be equivocal with respect to the timing of the flow responsible for the observed substrate. The same limitation applies to surface sediment samples whose grain size may be a response to the current at the time of sampling or some earlier flow event. Interpretations from grain size may be further hampered by changes in sediment supply and post-depositional processes such as bioturbation and dissolution. Benthic layers of turbid water, outlined in nephelometer profiles, provide a more immediate semi-quantitative measure

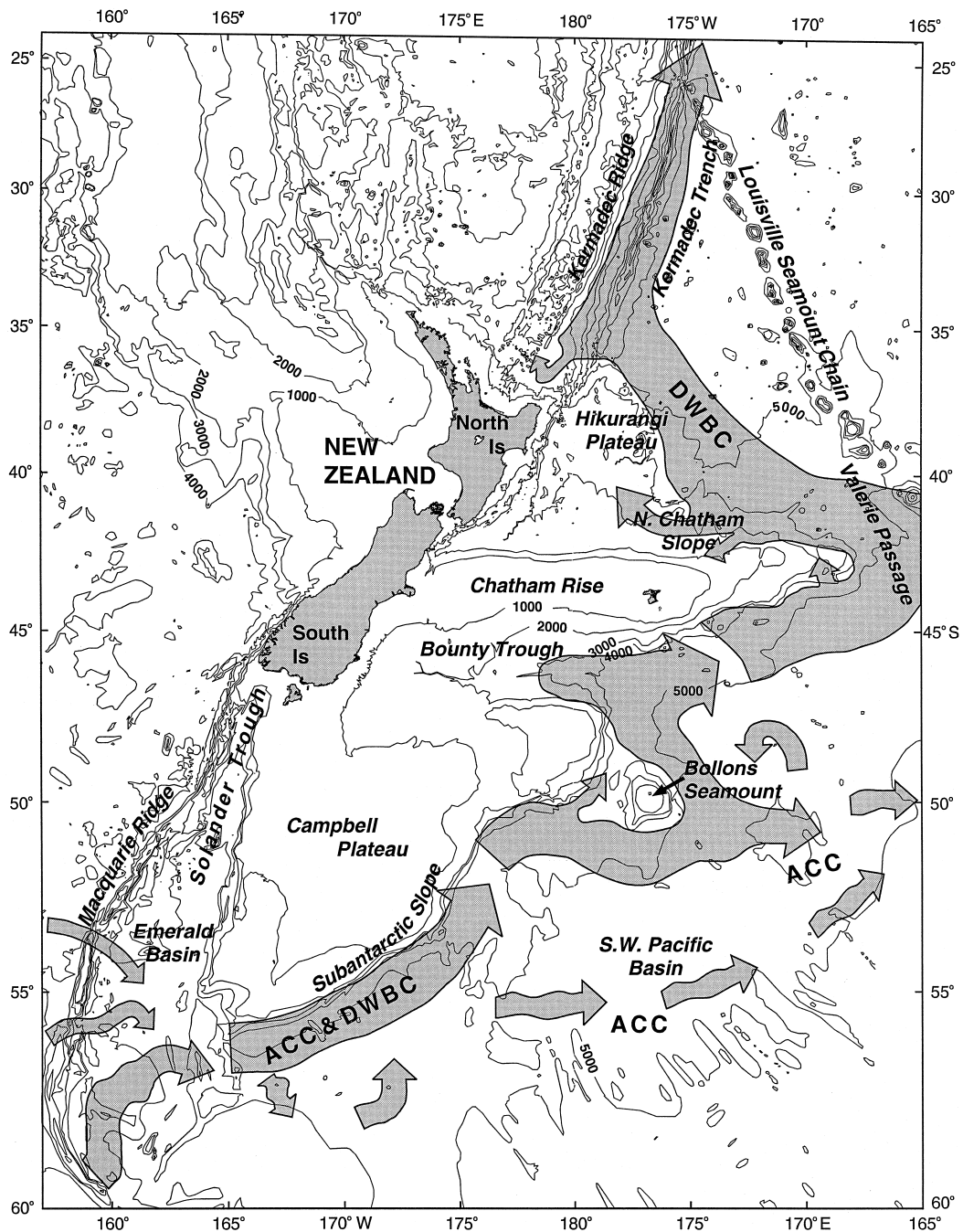


Fig. 1. Bathymetry in 1000 m contours together with generalised courses of the DWBC and ACC.

of the flow; the high turbidity usually equating with zones of rapid current and sediment resuspension (e.g., Eitrem et al., 1972). Interpretation of acousti-

cal responses and morphology recorded on high resolution seismic profiles provide some measure of current activity on the ocean floor (e.g., Damuth,

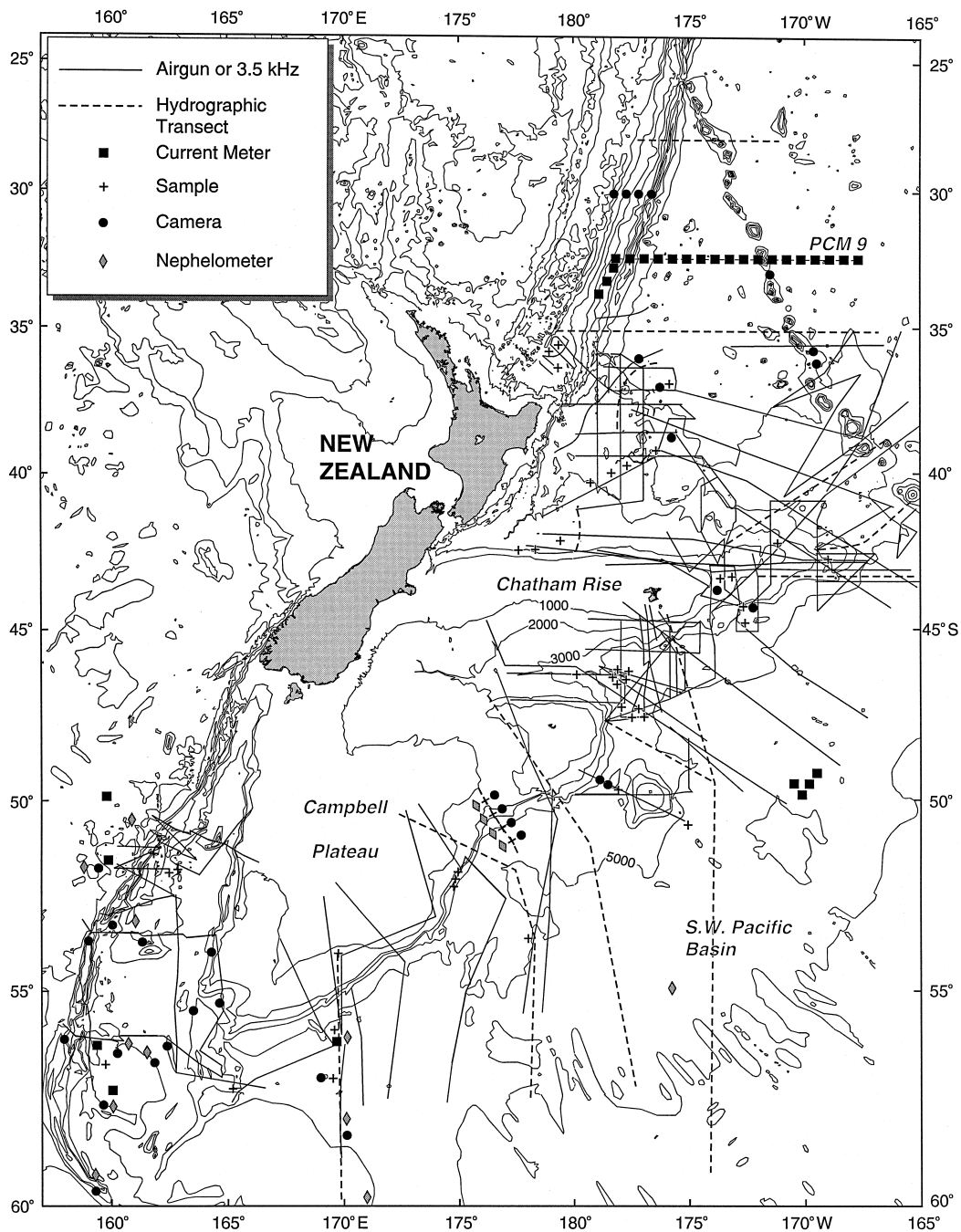


Fig. 2. Location of physical oceanographic and geological observational stations and transects including the World Ocean Circulation Experiment (WOCE) current meter array PCM-9.

1980; McCave and Carter, 1997). However, the timing of such activity is imprecise. Good quality 3.5

kHz records, for example, have a vertical resolution of ~ 0.5 m. With modern pelagic sedimentation

rates of ~ 4 cm/1000 year (data of Neil, 1991; Fenner et al., 1992) we can at best resolve ~ 12 ka of sedimentary record. In the case of unprocessed airgun records, such as those collected by *Eltanin*, the bubble pulse may mask the uppermost 20 m of seabed which equates to a potential loss of ~ 0.5 Ma of record. Even with these limitations, seismic profiles are useful because of their wide geographical coverage and their value as *generalised indicators* of long-term erosion and deposition.

Knowledge of the physical oceanography, in particular the geostrophic flow, stems from hydrographic transects off southernmost Campbell Plateau (Gordon, 1975; Mantyla and Reid, 1983); Bounty Trough (Walkington and Moore, 1995); Chatham Rise (Warren, 1973); Hikurangi Plateau (Nowlin et al., 1989; McCave and Carter, 1997) and Kermadec Ridge (Gilmour, 1973; Warren, 1981; Whitworth, 1994). Direct current measurements include short term (< 3 days) records from the southern Campbell Plateau (Jacobs et al., 1972, 1974). More substantive records come from beneath the ACC, east of Bounty Trough at $49^{\circ}40'S$ and $170^{\circ}30'W$ (Bryden and Heath, 1985), and from the DWBC at the WOCE line PCM-9 across the Kermadec Ridge and Trench at $32^{\circ}30'S$ (Fig. 2; Moore et al., 1993; Pillsbury et al., 1994; Whitworth, 1994).

2. Environmental setting of the inflow region

Deep waters from the Indian and Atlantic sectors of the Southern Ocean escape north into the New Zealand region via a series of fracture zones in the Southeast Indian Ridge (Rodman and Gordon, 1982). As these waters swing east, in consort with the ACC, the combined flow encounters the pronounced relief of Macquarie Ridge (Fig. 1). Deep seated gaps in the Ridge, north and south of Macquarie Island, allow current filaments to pass into the adjacent Emerald Basin (Gordon, 1972, 1975). However, the main body of the ACC passes around the end of Macquarie Ridge to cross Emerald Basin as a series of eddies and meanders (Boyer and Guala, 1972; Morrow et al., 1992). Not surprisingly, this is a region of high eddy kinetic energy (EKE) with eddies affecting depths to > 3000 m (e.g., Hollister and Nowell, 1991).

The perturbed flow continues east to the steep Subantarctic Slope bordering Campbell Plateau. This 3000–3500-m high western boundary steers the DWBC and ACC to the northeast, the flow being dominated by the ACC. En route, the ACC swings east to continue its circum-Antarctic journey (Bryden and Heath, 1985; Orsi et al., 1995; Belkin and Gordon, 1996) leaving the DWBC unescorted north of $49^{\circ}S$ (Carter and McCave, 1997). At this latitude the steep western boundary is replaced by the gently sloping floor of Bounty Trough resulting in a temporary reduction in current speed. Further north, re-establishment of a steep boundary at Chatham Rise, together with the constriction of Valerie Passage, cause the DWBC to intensify (Carter and McCave, 1994).

Once through Valerie Passage, the DWBC sweeps northwest along the subdued relief of North Chatham Slope and the steep flank of Hikurangi Plateau before taking a northerly course along the pronounced Kermadec Ridge (Carter and McCave, 1994). A second current filament appears to separate near Valerie Passage and bathe the eastern margin of the Louisville Seamount Chain. The filament eventually joins the main current at $26^{\circ}S$, where the seamount chain intercepts Kermadec Ridge (McCave and Carter, 1997).

3. Comparison of model output with observations

3.1. Circulation

The 10-year mean model circulation for selected depth levels is compared to geological and physical oceanographic data. For clarity and convenience comparisons are made for key physiographic regions along the current path, from Macquarie Ridge to Kermadec Ridge near the intersection with Louisville Seamount Chain (Fig. 1).

3.1.1. Macquarie Ridge–Emerald Basin

The LANL model generally confirms the strong perturbations caused by the interaction of the ACC with Macquarie Ridge as observed by Gordon (1972, 1975) and Rodman and Gordon (1982).

Strong jets pass through ridge gaps at 53°30'S and 56°S (Fig. 3a) causing a strong mean current across the floor of Emerald Basin (Fig. 3b). In nature, the jet pathway coincides with an extensively eroded basinal sediment cover bearing active bedforms and a dense, benthic nepheloid layer (Fig. 4; Heezen and Hollister, 1971; Carter and McCave, 1997). The model also has strong flows around the end of Macquarie Ridge at 59°S (Fig. 3) and some recirculation into Emerald Basin. This pattern approximates the meanders interpreted from hydrographic sections by Gordon (1972). In contrast, the circulation north of the meanders and zonal jets is weak as is reflected in the bioturbated muddy sediments and depleted nepheloid layers of northern Emerald Basin and Solander Trough (Fig. 4).

3.1.2. Campbell Plateau–Subantarctic Slope

From Emerald Basin, the ACC and DWBC continue along the Subantarctic Slope. Below 2475 m depth the modelled current is intensified along the slope (Fig. 3b). In nature this is manifested by (i) an extensive ferromanganese nodule and crust pavement on the floor of the Southwest Pacific Basin, and (ii) a scour moat along the base of the Subantarctic Slope (e.g., Glasby, 1976; Carter, 1989) (Fig. 4). The model abyssal flow is steered strongly by the 5000 m isobath. This, in turn, influences the overlying waters and causes part of the ACC to depart from the Subantarctic Slope and meander eastward at 55°S, 174°E (Fig. 3). Some of this separated flow recirculates southwestward in a gyre centred near 170°E off southern Campbell Plateau; this gyre apparently being captured in a CTD section along that meridian (Gordon, 1975). The presence of an eastward meander may be supported by geological data. Sediment isopachs, proximal to the base of the Subantarctic Slope, exhibit a localised, northeast-aligned lobe that may be caused by the eastward veering of the flow near 55°S (Fig. 4). In a review of hydrographic sections, Orsi et al. (1995) identify a frontal feature at ~55°S as the signature of the Subantarctic Front which is usually considered to be the northernmost zonal jet of the ACC (e.g., Patterson and Whitworth, 1990). However, Heath (1981) and Belkin and Gordon (1996) consider the Front to be the boundary flow that continues northeastward along Campbell

Plateau. The model favours the latter interpretation, and shows a zone of strong eddying flow separating from the Subantarctic Slope at 49°S. Meanders and recirculation within this zone are evident at 2475 m between 175°W and 165°W, and extend almost unattenuated to the seafloor as corroborated by the current measurements of Bryden and Heath (1985).

3.1.3. Bounty Trough

Departure of the ACC from the Subantarctic Slope at 55°S and 49°S, coupled with a marked reduction of western boundary relief at Bounty Trough, argues for a reduction in the abyssal velocity field. This contention is borne out by (i) a weak circulation about the Trough mouth (Fig. 5), and (ii) the growth pattern of the Bounty Fan which extends directly across the anticipated path of the DWBC (Carter and Carter, 1996) (Fig. 4). This pattern contrasts with that of fans deposited under strong flows, the tendency being for an alignment of fan morphology with the current (e.g., Schneider and Heezen, 1966; Carter and McCave, 1994). Nevertheless, Bounty Fan may still show a sedimentological response to the DWBC. One possible line of evidence is a coarsening of sediment into the path of the DWBC, although this trend may also reflect normal fan sedimentation (Carter and McCave, 1997).

The southern entrance to Bounty Trough is guarded by Bollons Seamount which rises from 4500 m to ca. 900 m depth (Fig. 1). Although the model fails to register the seamount until 3500 m depth, it shows the edifice diverting the flow to the east with little recirculation into Bounty Trough. In contrast, seismic data suggest that part of the DWBC recirculates in the outer reaches of the Trough where it is steered by a prominent sill formed across the trough axis at 3500–4000 m depth (Carter and McCave, 1997) (Fig. 4). The sill has an eroded sediment cover which is consistent with flow intensification against this small but prominent boundary.

3.1.4. Eastern Chatham Rise–Valerie Passage

North of Bounty Trough, the bathymetry is dominated by the 3000–4000-m high flanks of eastern Chatham Rise. For depths < 2500 m, the model has the flow moving south around the Rise and into

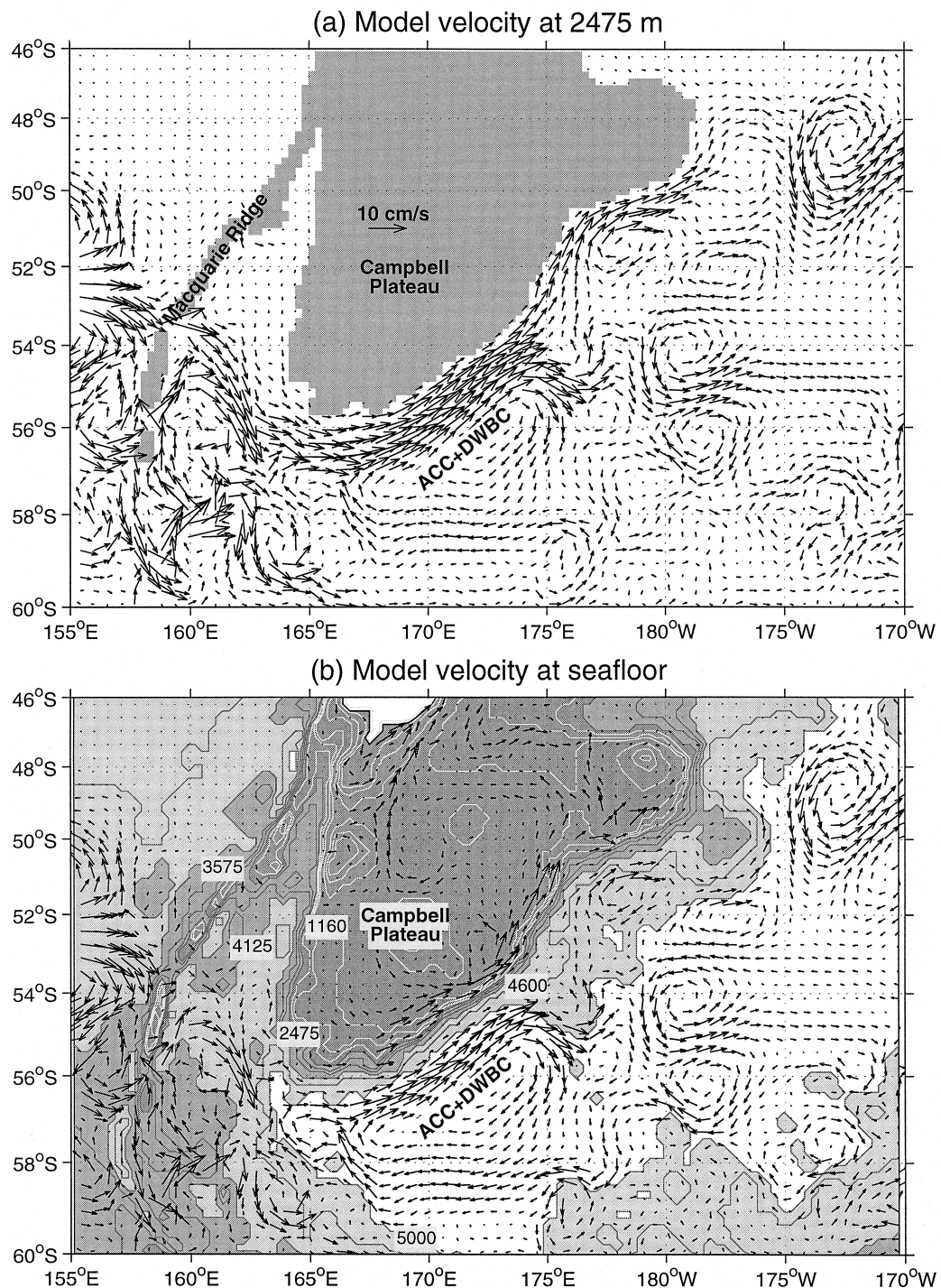


Fig. 3. The LANL model circulation south of Chatham Rise. Vectors are plotted for every second model grid point. (a) Mean velocity at depth 2475 m. (b) Mean velocity at the seafloor plotted over model bathymetry with shading changes at 2475, 4125 and 5000 m, depths. Other model depths are contoured. The 10 cm/s scale vector shown in (a) also applies to (b). ACC + DWBC is the combined flow of the ACC and the DWBC.

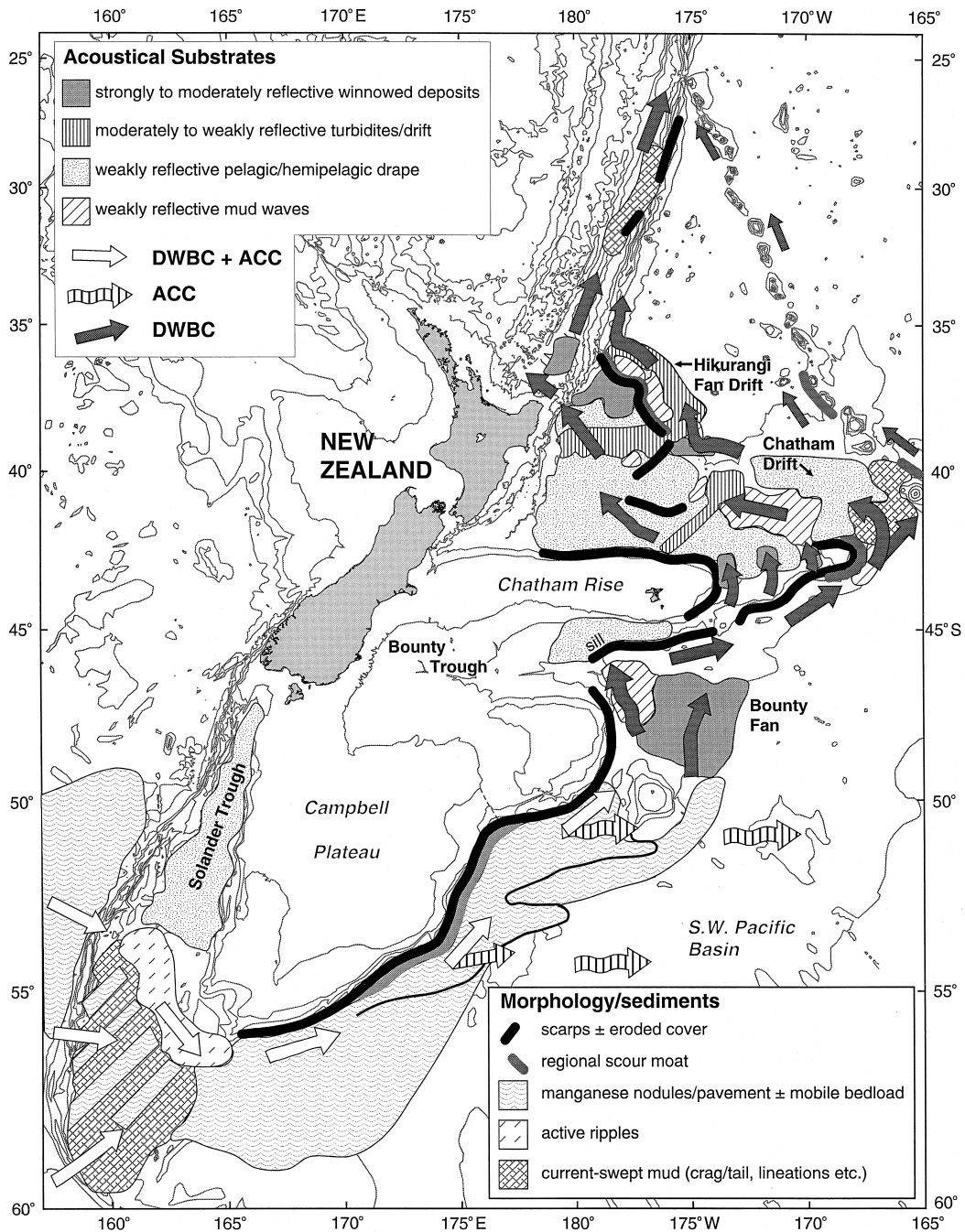


Fig. 4. A generalised summary of (a) photographic and sampling observations and (b) acoustical substrates interpreted from 3.5 kHz profiles (Carter and Carter, 1996; McCave and Carter, 1997). These data assisted identification of the abyssal circulation, outlined by the arrows. Strongly to moderately reflective deposits occur mainly within topographically intensified parts of the DWBC and where this current is enhanced by the ACC. Weakly reflective pelagite/hemipelagite zones presently encounter tranquil flow; an interpretation confirmed by the prevalence of bioturbated, muddy substrates. Observations of the turbidite/drift substrates also suggest modern tranquil flow, although this is obviously interrupted by turbidity currents with frequencies of several centuries or longer.

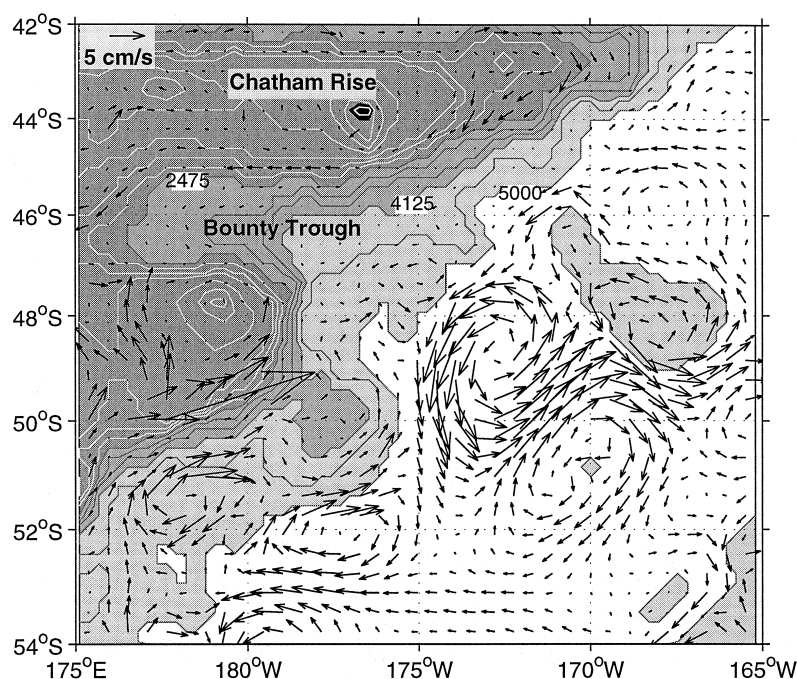


Fig. 5. LANL model velocities at the ocean floor off Bounty Trough depicting weak and strong flows within and outside Bounty Trough, respectively. Vectors are plotted over model bathymetry, with shading changes at 2475, 4125 and 5000 m. Other depths are contoured. A 5 cm/s scale vector is shown at top left.

Bounty Trough, whereas at greater depths the current direction is reversed (Fig. 6). Limited support for this reversal in flow is provided by potential density sections southeast and east of Chatham Rise (Warren, 1973; unpublished data from R.V. *Knorr*, 1973 survey). The derived geostrophic fields indicate a southward current adjacent to the western boundary at 2500–3200 m. In deeper water, the flow switches mainly to the north, although there may be fluctuations relating to the eddy field of the nearby ACC.

As the DWBC approaches Valerie Passage the model flow splits into (i) a northward, relatively shallower arm extending through the Passage, and (ii) a deeper arm directed east to the Louisville Seamount Chain (Fig. 6). Observations are insufficient to corroborate the split, but in Valerie Passage the geological data support an intensification predicted by the model for depths > 2500 m. Certainly, the eastern flank of Chatham Rise is locally scoured (Fig. 4). Furthermore, drifts on the floor of Valerie Passage—indicators of long term deposition—have signs of recent erosion such as truncated seismic

reflectors, extensively furrowed surfaces, and a well-defined benthic nepheloid layer (McCave and Carter, 1997). The model current reduces towards the eastern side of Valerie Passage—a trend that is matched by the appearance of uneroded pelagic drape on the Passage floor.

Returning to eastern Chatham Rise, the model grid size is too large to discern three gaps (9–15 km wide; 90–130 km long) that traverse the Rise crest at depths between 1750–2900 m (Carter and McCave, 1994). The distribution of acoustical substrate types, derived from 3.5 kHz records, suggests that bottom scouring currents pass northwards over the Rise (Fig. 4). Assuming a mean current speed of 5 cm/s over a total cross-sectional area 11 km² for the gaps, then the potential volume transport could be 0.5×10^6 m³/s or ~5% of the total DWBC transport.

3.1.5. Louisville Seamount Chain

As noted earlier, the model circulation splits near Valerie Passage with an arm extending along the eastern margin of Louisville Seamount Chain. The

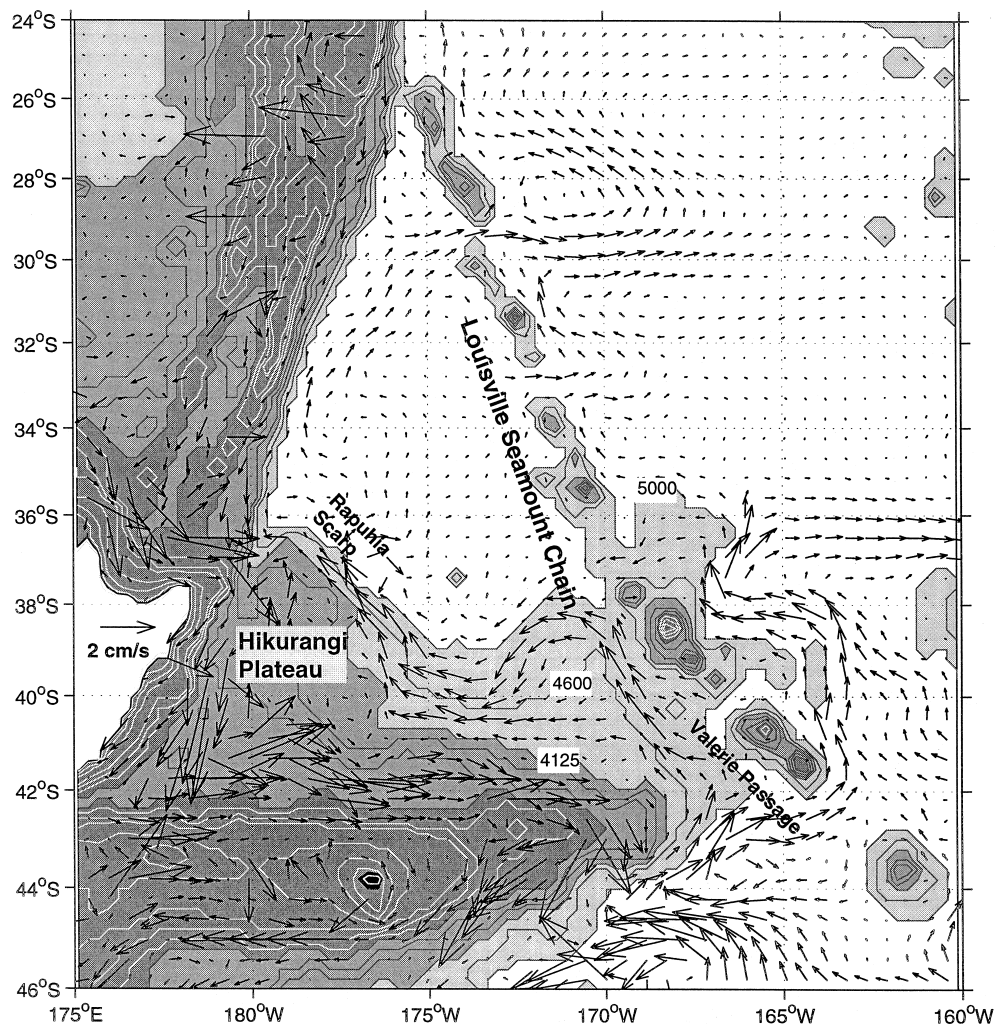


Fig. 6. LANL circulation for the ocean floor north of Chatham Rise superimposed on the model bathymetry. A 2 cm/s scale vector is shown at mid left.

existence of this DWBC arm has limited support from a series of seismic transects displaying preferential erosion or non-deposition on the eastern flanks of seamounts between 44–40°S (Watts and Weissel, 1988; McCave and Carter, 1997). By comparison, erosion or non-deposition affects the western side of the seamount chain between 38°–36°S. In the apparent absence of strong flows in this sector, the scouring could be a local effect caused by currents leaking through gaps in the chain. North of 33°S, current velocities recorded at WOCE PCM-9 line (M. Moore;

personal communication, 1996), a seismic profile at 31°30'S (McCave and Carter, 1997), and a geostrophic section along 28°S (Warren, 1973) suggest re-establishment of a stronger current along the eastern side of the chain. Similarly, the model portrays a boundary effect along the same reach, although this effect may be a response to spurious eastward migration of the main DWBC from the Kermadec Ridge (see Section 3.1.7).

The LANL model may underestimate the boundary flow along Louisville Seamount Chain because

of weak bathymetric control. At the 2000 m level, between 44°S and 25.7°S, the model depicts only two volcanoes in the 2300 km long chain whereas 40 minor and major volcanic peaks actually reach that depth (Lonsdale, 1988). At the 3000 m level, eight model peaks are evident whereas in nature 45 peaks occur and account for ~40% of the chain length (cf. Figs. 4 and 6). This underestimation of a significant but porous western boundary reflects the model grid size which is larger than most seamount peaks. Conversely, the model exaggerates the size of the gaps between volcanoes and therefore may underestimate the velocity of intergap currents. For example, the average model gap at the 3000 m level is ~250 km wide but the actual average gap is only ~60 km wide. This difference potentially leads to flow through the model gaps having greater total transport yet weaker peak velocities than occur in reality. Thus, while the model has a weak circulation in many of its gaps, the real situation, as portrayed in bottom photographs, is a current that periodically strengthens to erode the substrate (McCave and Carter, 1997).

3.1.6. North Chatham Rise–Hikurangi Plateau

Once through Valerie Passage, the main DWBC interacts with a variable topography causing vertical and horizontal separation of the flow. West of the Passage, in depths <2500 m, the prevailing model circulation is decidedly eastward along Chatham Rise—a condition supported by a scattering of hydrographic measurements (Heath, 1981; Chiswell, 1994; Tomczak and Godfrey, 1994). Below 2500 m the model anticipates a ‘zone of no motion’, for at 3000 m the circulation is directed westward as the upper part of the DWBC moves onto the gently inclined North Chatham Slope and the 3500–4000-m deep Hikurangi Plateau (Fig. 6). The upper DWBC is shown as weak presumably because of an insignificant western boundary. Certainly, this is the case in nature where the most common elements of relief are only scattered seamounts (Carter and McCave, 1994). More direct evidence of a weak circulation is the prevalence of uneroded, fine-grained sediments including pelagic ooze and muddy overbank deposits from Hikurangi Channel (Lewis, 1994). Localised scouring of plateau sediments occurs where the upper DWBC is constricted between seamounts, but

this process is at a scale too small for the model resolution. Subtle changes in sediment texture also support the model. Modes in the terrigenous silt fraction, and the silt/clay ratio increase irregularly eastward into the zones of strong flow that characterise the lower part (>3500 m) of the DWBC (McCave and Carter, 1997).

Immediately northwest of Valerie Passage, the model has the lower DWBC passing along the 4400 m isobath to form a large anticlockwise loop about 39°S; 171°W (Fig. 6). This circuitous course is surprising as there is virtually no relief at that locality. The validity of that particular pathway is also challenged by a geostrophic transect near the loop (Fig. 2; Nowlin et al., 1989). Derived velocities fail to show a DWBC core at 5000 m. Instead, the transect displays fairly uniform velocities of 1.5–2.0 cm/s to the northwest with a slight intensification against the foot of North Chatham Slope.

The loop ends against Rekohu Drift off southern Hikurangi Plateau where the topography steers an accelerating DWBC to the northwest (Fig. 4). The modelled intensification is confirmed by geological data. Initially the current is guided by the gentle flanks of Rekohu Drift but later intensifies against the 1200-m high, 300-km long wall of Rapuhia Scarp. There, 3.5 kHz profiles outline a 170-m deep (maximum) boundary channel scoured along the full length of the scarp base (Fig. 4). Sediments on the channel floor contain sand and gravel with mobile bedforms, and are overlain by a thick nepheloid layer. Thus, the DWBC is sufficiently strong to induce erosion, albeit along a narrow channel <5 km wide. Immediately east of the channel the model points to flow deceleration—a change supported by the deposition of Hikurangi Fan Drift along full length of Rapuhia Scarp (Fig. 4). A preponderance of muddy sediments, devoid of current induced structures, together with a reduced benthic nepheloid layer further indicate a tranquil hydraulic regime (McCave and Carter, 1997). A reduction in speed of the model DWBC near 36°S coincides with its departure from Rapuhia Scarp and subsequent passage across 7000-m deep head of Kermadec Trench.

3.1.7. Kermadec Ridge

The final New Zealand leg of the DWBC extends 1700 km along the imposing western boundary of

Kermadec Ridge (Figs. 4 and 6). According to the model, the flow is strongly bi-directional. For levels shallower than 3500 m a topographically intensified current passes south along the full length of the Ridge (Fig. 7), whereas deeper levels display a weak, northward boundary current.

The relative strength of the north and south flows is at variance with the majority of observations in the region. Hydrographic sections at 35°S, 32°30'S and 28°S all yield significant geostrophic flows to the north for depths occupied by the DWBC. For instance, Warren (1973) calculated a velocity of up to 18 cm/s northward at 5000 m in a section at 28°S. The model mean velocity at this section (Fig. 7) is no more than 0.3 cm/s northward at 5000 m near the ridge, being overwhelmed by the southward flow with a maximum of 3 cm/s at 2500 m depth. This southward intermediate depth flow can be traced back to a boundary current that enters through the Samoa Gap, and is supplemented by a westward inflow at 15°S. There is observational support for this sense to the boundary current in the analysis of

South Pacific hydrography and tracer distributions by Reid (1986). He identified a southward current, at 2500 m along Kermadec Ridge, supplied by westward flow near 15°S. Reid's data fail to reconcile the flow as a boundary current, but the current meter records from WOCE PCM-9 show a mean southward flow of 1 cm/s at 2500 m within 100 km of the ridge. However, the magnitude of the southward boundary current, compared to the deeper northward DWBC, appears underestimated by the model as nowhere along Kermadec Ridge does the model achieve northward velocities approaching observations.

A further discrepancy is that at 29°S. There, the model abyssal current swings east to pass through a gap in the Louisville Seamount Chain before resuming a northward course to eventually intercept Kermadec Ridge at ~22°S (Fig. 6). Hydrographic sections at 28°S (Warren, 1973) and 22°S (Gilmour, 1973) have the DWBC against Kermadec Ridge implying continuity to the north without an eastward deviation.

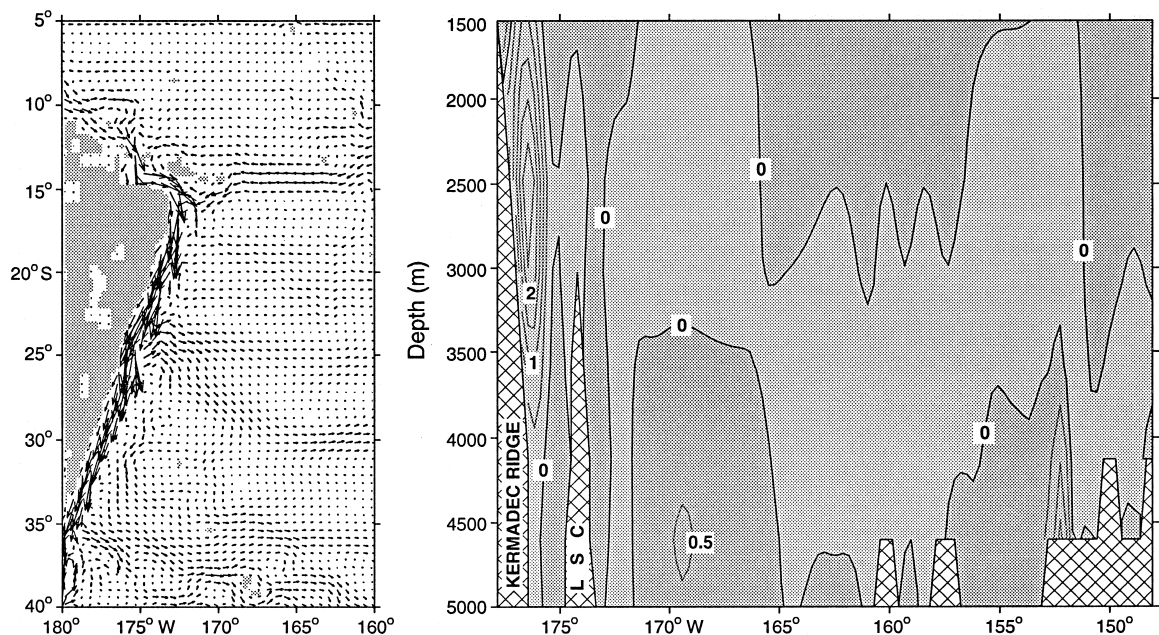


Fig. 7. Right: The modelled velocity at latitude 28°S, with light shade being the flow to the south and dark shade the northward current. LSC = Louisville Seamount Chain. Left: Modelled circulation at 2475 m showing the pronounced southward flow from the Samoa Passage along the Tonga–Kermadec Ridge. North (east) component of the vectors is plotted at a scale of 0.6° latitude (longitude) per cm/s.

Geological information sheds little light on flow direction along Kermadec Ridge, but has some bearing on current intensity. Current induced structures such as crag and tail, and a well developed benthic nepheloid layer confirm the general strength of the DWBC as portrayed in the model levels > 3500 m (McCave and Carter, 1997). Furthermore, these data suggest that the zone of maximum flow is between 5000–6000 m depth which also agrees with the level vector plots.

3.2. EKE

A further test can be made through a comparison of modelled abyssal EKE with various geological parameters (cf. Fig. 4 with Figs. 8 and 9). A similar but more generalised analysis was made by Hollister and McCave (1984). These authors related data from bottom photographs with the variability of sea surface height—a parameter that compares well with EKE (Schmitz, 1984).

The highest EKE at the seafloor, $80 \text{ cm}^2/\text{s}^2$, occurs off southernmost Campbell Plateau (Fig. 8). Here, the sediment cover is extensively eroded and ferromanganese nodules and crusts are pervasive (Goodell et al., 1973; Glasby, 1976). The pattern of eddy energy is matched by changes in the character of the ocean floor. For example, in Emerald Basin energy values range from $10\text{--}30 \text{ cm}^2/\text{s}^2$ and the substrate supports a band of active bedforms bordered by current-swept mud. By comparison, the seabed along the foot of northwestern Campbell Plateau is mantled mainly by ferromanganese deposits that extend as far as Bollons Seamount approximating the trend of the $10 \text{ cm}^2/\text{s}^2$ isolines (Carter, 1989). Although similar to nodules and crusts in the zone of highest EKE, the northwestern deposits have a thin veneer of sediment that suggests either an intermittent or less rigorous current regime (Carter and McCave, 1997).

A second zone of high EKE occurs where the ACC separates at 49°S and forms an intense, deep-

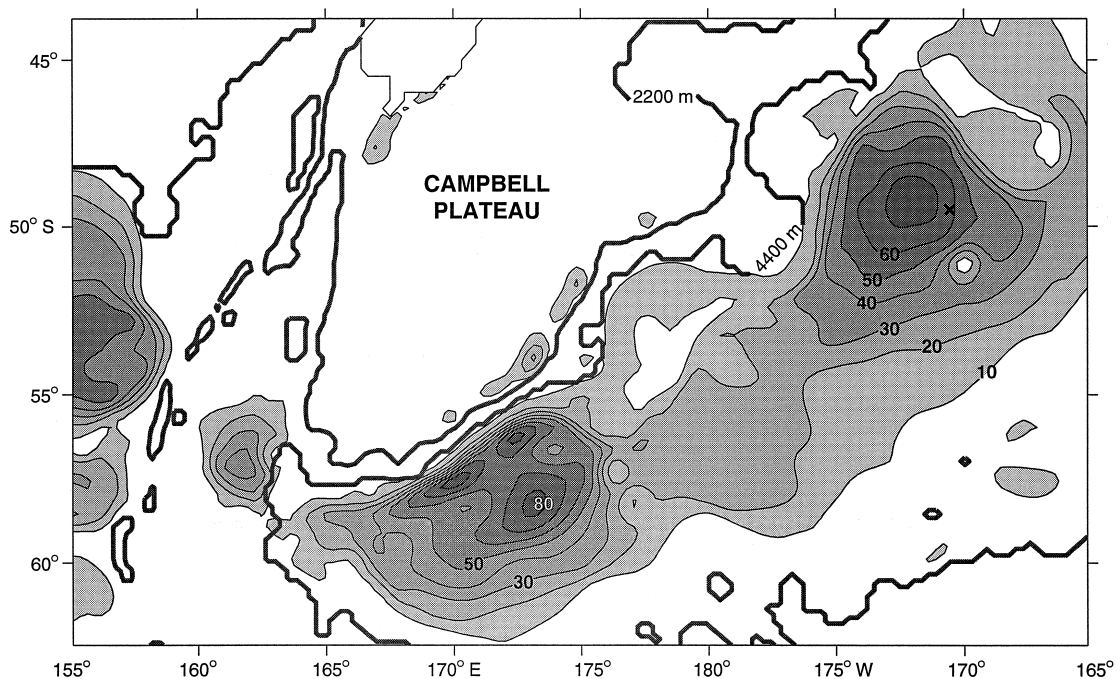


Fig. 8. Model EKE at the ocean floor south of Chatham Rise where the flow is strongly influenced by the ACC. Cross at 49° is the location of the moorings of Bryden and Heath (1985). Values are in cm^2/s^2 .

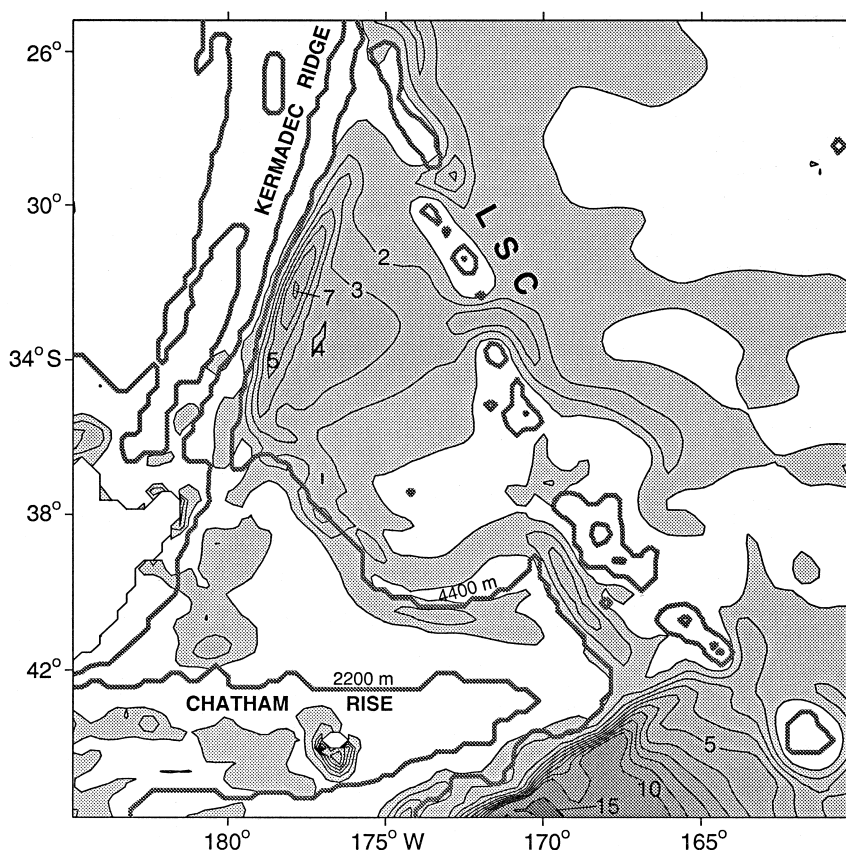


Fig. 9. Model kinetic energy at the seafloor for the region north of Chatham Rise. Values are in cm^2/s^2 .

reaching eddy centred on 172°W (Fig. 3a and Fig. 5). The current meter array deployed by Bryden and Heath (1985) is located in this feature and allows us to compare direct estimates of observed mean current and EKE with the model. The mean velocity is modelled well (Table 1) with magnitudes agreeing to within 25% and directions to within 10° . Given the

discrepancies between observations and the model identified previously, the agreement in Table 1 may be fortuitous. However, within several grid points of the mooring location the model velocity is little changed (Fig. 5). If the horizontal scale of the eddy is equally large in reality, this would favour agreement because a lateral offset of the position of flow

Table 1

Comparison of mean east (u) and north (v) velocity components and EKE values from Bryden and Heath (1985) mooring P ($49^\circ24'\text{S}$, $170^\circ30'\text{W}$) with the nearest LANL model grid point

Depth (m)		u (cm/s)		v (cm/s)		EKE (cm^2/s^2)	
Current meter observation	Nearest model level	Obs.	Model	Obs.	Model	Obs.	Model
1000	1160	8.06	10.8	3.05	4.40	169.2	194.3
2000	1975	4.17	6.86	3.12	3.40	58.4	94.2
4000	4125	4.10	4.95	3.67	2.96	38.1	64.1

features would be less likely to play a confounding role in the comparison. It is therefore likely that this favourable comparison is a robust result. Mean velocities agree best at depth, but here the model overestimates the EKE by as much as 40%. Comparisons with satellite sea surface height variability (McClean et al., 1997) have shown the LANL model tends to overestimate surface EKE in several sectors of the ACC including the New Zealand region. Elevated values at depth therefore do not necessarily indicate that the model fails to vertically attenuate eddy energy in an adequate way. Many more mooring comparisons would be required to address this issue.

North of the 49°S separation, the model shows a marked reduction in abyssal EKE—a change that is not matched by observations. Locally elevated model values of 15–20 cm²/s² show up at the base of Chatham Rise (Fig. 9) where the DWBC becomes distinct from the ACC and advances toward the bifurcation near Valerie Passage. However, subsequent local peak values reach no more than 4 cm²/s² in the Passage itself. Similar values are derived for the base of Rapuhia Scarp. But there, an intensified DWBC is actively scouring the substrate to form transverse sand ripples (Carter and McCave, 1994; McCave and Carter, 1997). These bed responses to the current are similar to those in southern Emerald Basin where energy exceeds 40 cm²/s² (Fig. 8). Similarly, photographic transects down Kermadec Ridge point to periodic flows capable of scouring moats around gravel clasts, forming crag and tail structures, and generally smoothing the bed (McCave and Carter, 1997). Yet, model energy is only ~8 cm²/s². Either the model is not registering these narrow zones of active scour or it is underestimating the EKE in the DWBC outside the influence of the ACC.

The latter case has validity according to a recent analysis of LANL model output and current meter data from the WOCE PCM-9 mooring (Moore and Wilkin, 1998). The comparison revealed similarities between observed and modelled current variability on the open ocean floor, but marked discrepancies close to Kermadec Ridge. Moore and Wilkin considered the components of velocity separately, being interested in the polarisation of velocity due to planetary wave dynamics. Accordingly, we have indepen-

dently calculated EKE for the current meters closest to the seafloor from the WOCE PCM-9 data (Pillsbury et al., 1994). At 2530 m depth, on the steep flank of Kermadec Ridge, EKE was 84.8 cm²/s². It decreased from that maximum value to 17.8 cm²/s² at 3880 m, 12.5 cm²/s² at 5134 m, and 6.5 cm²/s² at the foot of the Kermadec Ridge in 5900 m depth. In contrast, the model EKE is < 9 cm²/s² at any level below 2000 m depth. A similar discrepancy appears in the mean, with the model transport of waters deeper than 3000 m being some 5 times lower than observed.

Moore and Wilkin (1998) reveal the presence of bottom-trapped topographic waves generated by the impact of planetary Rossby waves on Kermadec Ridge. They suggest the discrepancy in EKE is due partly to a poor simulation of these bottom-trapped waves. Strong boundary friction in the model weakens the mean current immediately adjacent to the Ridge, and this also affects the wave-induced variability. Whether too much friction can account for the weak overall transport of the DWBC is uncertain. It is possible that the anomalously strong southward intermediate depth flow along the Kermadec Ridge partially negates the northward DWBC.

If the principal source of current variability is the interaction of westward propagating Rossby waves with the bathymetry, then this would account for other regions where the geology indicates elevated variability such as along Rapuhia Scarp, the eastern side of Louisville Seamount Chain, and even in Valerie Passage if the seamount chain fails to fully block the westward passage of Rossby waves. The comparison with WOCE PCM-9 would suggest that the model significantly underestimates abyssal EKE in these regions also.

4. Discussion

4.1. Topography

Model topography departs from nature in some localised areas to produce spurious circulation patterns. One conspicuous example is the eastward deviation of the modelled DWBC from Kermadec Ridge

to a gap in the Louisville Seamount Chain at 29°S (Fig. 6). In contrast, hydrographic sections indicate flow continuity along the ridge to at least 22°S. This deviation is a reflection of the model topography which artificially blocks the DWBC at the 5000 m level where Louisville Seamount Chain collides with Kermadec Ridge. To the contrary, the ridge and seamount chain are separated by a 26-km wide passage deeper than 5000 m (Fig. 1; Ballance et al., 1989; CANZ, 1997). Hence, some northward continuity of the boundary current is assured. Nevertheless, the passage is narrow compared to the overall size of the DWBC. Thus, some reflection of the current may take place as suggested by a southward return flow along the western side of Louisville Seamount Chain in the hydrographic sections at 32°30'S and 28°S (M. Moore, unpublished data for WOCE line PCM9; Warren, 1973). Even if some eastward deviation of the DWBC occurs, the model markedly overestimates the size of the 15-km wide gap in the Seamount Chain at 29°S. Therefore, any passage through the real chain would be restricted and possibly deflect any eastward current. In fact, the model overestimates the size of most gaps in the Louisville Seamount Chain as well as underestimating seamount relief. Hence, the real chain is a more effective western boundary than its model counterpart.

Bollons Seamount is another feature with underestimated relief. The model seamount is about 1000 m above the ocean floor whereas the actual relief is 3250 m. Therefore, Bollons Seamount presents a significant boundary to the ACC/DWBC. This edifice may assist the eastward deflection of the zonal jet marking the northern limit of the ACC in the region (e.g., Bryden and Heath, 1985). Also, the model has not captured the 27-km wide gap at 4500 m depth between Bollons Seamount and Campbell Plateau. This gap allows some of the DWBC to pass into Bounty Trough (Carter and McCave, 1997). By comparison, the model has the gap closed at 3850 m and the abyssal flow is re-routed east around the seamount (Fig. 5).

The LANL model is sensitive to topography, a trait that is common to other models of this type (e.g., Grose et al., 1995). Currents are forced to follow deep isobaths even though in nature these contours may outline regions of low relief, e.g., the

5000 m isobath off (i) Campbell Plateau at about 175°E, (ii) north of Valerie Passage at 40°–35°S and (iii) along the eastern side of Louisville Seamount Chain. Adherence to a contour is a response to the step-like topography of the model. For instance, the 5000 m depth level is marked by a 400 m high step, which guides the current to conserve barotropic vorticity (e.g., Grose et al., 1995). Hence, the influence of the step is felt for almost the full depth range of the DWBC resulting in a deceptive circulation pattern. An example is the anticlockwise circulation at the exit of the Valerie Passage (39°S, 172°W) where the current is guided by the step separating the 4400 m and 4800 m isobaths. In reality the bathymetry is the gently inclined foot of the North Chatham Slope which exerts little topographic control.

4.2. Model resolution

Even though the LANL model has one of the smallest grid sizes of global circulation models, it still misses topographic features that can significantly affect the abyssal circulation. The blockage of the DWBC path along Kermadec Ridge by the Louisville Seamount Chain and consequent diversion of the flow along 29°S (Fig. 6) may not only be a function of bathymetric database used by the model, but also may reflect the inability of the model to resolve the 26-km wide passage separating the ridge from the seamount chain. The same comment applies to the 27-km wide gap between Bollons Seamount and Campbell Plateau.

Another set of unresolved, but potentially influential bottom features is the north–south gaps across the crest of eastern Chatham Rise. There, up to 5% of the DWBC may be directed *across* the Rise whereas the model shows the current moving *around* the Rise (Fig. 4). A similar situation affects the uppermost DWBC at the southern end of Kermadec Ridge. The model fails to resolve several narrow gaps in the Ridge and therefore directs the flow along the Ridge, albeit in a direction opposed to the observed current. In the natural world, some of the DWBC leaks through the 2000–2250-m deep gaps with an observed total volume transport of 0.2×10^6 m³/s (Warren et al., 1994).

4.3. EKE

Despite potential differences between the temporal scales of geological and physical oceanographic observations, there is broad agreement between modelled EKE and flow conditions interpreted from geological features (cf. Figs. 4 and 6). However, the model clearly achieves better simulation of abyssal EKE in the Southern Ocean compared to the SW Pacific Ocean, north of Chatham Rise. Variability in the south is dominated by instability processes within the ACC. This variability penetrates deeply, driving abyssal current variations that shape the sediment cover. In the northern abyssal region, the simulation of mean DWBC transport and variability is poorer. If Moore and Wilkin (1998) are correct, namely that abyssal variability is driven by mid-latitude Rossby waves and their interaction with bathymetry, then it appears the model would benefit from an improved representation of these processes.

5. Conclusions

Physical oceanographic observations of the abyssal circulation in the New Zealand region are few and widely scattered especially in the Southern Ocean. In that light, the LANL model fulfils a key role in providing a broad perspective of the abyssal circulation. The introduction of sedimentary and geophysical information to the oceanographic database has improved our capability to test the model circulation. Results are encouraging even though some of the geophysical data may represent current effects that occurred at some unspecified time in the past millennia.

Broad agreement exists between the modelled and observed abyssal circulation patterns. Zones of strong predicted flows, in particular the ACC-affected region south of Chatham Rise, correlate well with active sediment transport, well-developed nepheloid layers, and eroded sediment cover. Furthermore, the competency of the current, as indicated by the modelled EKE, relates reasonably well to seabed characteristics such as bedform types and size of the nepheloid layer.

North of Chatham Rise correlations are less clear. Observations confirm current directions near pro-

nounced western boundaries such as Rapuhia Scarp and Kermadec Ridge. However, the intensity and variability of the boundary current, as portrayed by EKE, are markedly less than that indicated by the sediments. Such low EKE values may stem from an underestimation of both mean transport and variability within the model DWBC. For areas with a subdued western boundary, the model has the DWBC closely following isobaths even though in nature these contours occupy zones of almost flat topography. This adherence to isobaths is a response to the model bathymetry being represented as a series of discrete steps.

Further artificialities are created by the model's bathymetric database and its inability to resolve small but important bathymetric features. Of note is the artificial joining of Kermadec Ridge and Louisville Seamount Chain. This merger induces a major eastward diversion in the DWBC whereas in the reality the ridge and seamount chain are separated by a narrow passage that allows northward continuity of the flow. Appropriate changes in model bathymetry and reduced friction should lead to a more realistic transport for the DWBC. Further analysis is also required to determine why the model exhibits strong reverse flow at intermediate depths along Kermadec Ridge; this flow appears to negate much of the known northwards DWBC transport.

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